Groundwater storage – streamflow relations during winter in a subarctic wetland, Saskatchewan

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Wetland drainage systems are shown to be hydrologically active during winter. Water storage in various terrain types changed over the winter as a result of intrabasin transfers between terrain types, primarily from outlying mineral terrains to centrally located groundwater controlled wetlands, and due to winter streamflow. Mineral terrain and bog lost 97 and 25 mm of water, respectively, whereas fens gained 28–51 mm. A water balance indicated that mineral terrain yielded almost twice as much water as was released as streamflow, and that much of this excess was being stored in the fens where groundwater seepage at the surface resulted in icings. Bogs had little ability to sustain winter streamflow. Diminishing streamflow in early winter coincided with freezing of the surface layers of peat, which normally transmit most of the water. However, streamflow was maintained throughout winter by water transmitted through the fens.

La présente étude démontre que les systèmes de drainage des terres inondables demeurent hydrologiquement actifs durant la saison d'hiver. Le stockage de l'eau dans des terrains de types différents est modifié durant l'hiver à cause des transferts intra-bassins entre les divers types de terrains, allant principalement des terrains vaseux externes vers le centre des terrains inondables qui sont influencés localement par les eaux souterraines, et aussi à cause d'un certain écoulement fluvial en hiver. Le terrain vaseux et la tourbière ont perdu 97 et 25 mm d'eau, respectivement, tandis que les marécages ont gagné de 28 à 51 mm. L'équilibre hydrique a révélé que le terrain vaseux fournissait presque le double de la quantité d'eau évacuée par écoulement fluvial, et que la majeure partie de ce surplus d'eau était emmagasinée dans les marécages où le suintement des eaux souterraines à la surface prenait en glace. Les tourbières n'offraient qu'un faible potentiel de rétention des eaux d'écoulement fluvial durant l'hiver. La diminution de l'écoulement fluvial au début de l'hiver coïncidait avec le gel des couches à la surface de la tourbe qui normalement transmettait la majeure portion de l'eau. Cependant, l'écoulement fluvial était entretenu durant tout l'hiver grâce à une alimentation en eau par les marécages.

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Introduction

In temperate and subarctic regions, winter conditions dominate for 3–6 months of the year, but few process studies are undertaken because it is often assumed that basins are hydrologically inactive at this time. In wetland systems, the flow-generating processes typically occur at or near the surface (Bay 1969; FitzGibbon 1982; Roulet and Woo 1986), so they are strongly affected by freezing. In winter, the surface progressively freezes and thereby impedes hydrological activity in the acrotelm, which is the most permeable upper layer of peat (Ingram 1978). However, the lower, less permeable layer (the catotelm) remains unfrozen and retains a limited capacity for water transfer. This affects the water storage function of the various terrain types and the release of stored water for streamflow.

Not all wetland types have similar responses to hydrologic stress. Bogs are oligotrophic wetlands, receiving nutrients and water only from meteoric sources, whereas fens are groundwater controlled eutrophic wetlands that receive mineratrophic waters. Bay (1969) observed that runoff from a small bog ceased in winter, whereas a local fen produced flow throughout. Bavina (1972) reported that fens normally lost more water in winter than they gained, although continuous groundwater input to fens replenished storage, which sustained streamflow. Schwartz and Milne-Home (1982) found that stream chemistry indicated that winter runoff from wetlands originated in mineral terrains.

Previous studies did not involve detailed observations of hydrological processes during winter, or else specific data were unavailable (e.g., Bavina 1972). However, Price and FitzGibbon (1982) showed that when saturated peat in fens froze, it formed an impermeable concrete frost layer that confined incoming groundwater and caused artesian pressure to develop. In mineral terrain and bog, they found that the water table was below the freezing front in winter. Porous granular frost formed at the surface, but the zone primarily responsible for groundwater drainage remained unaffected by it. The effect of freezing on the basin water transfer in wetlands has not been previously studied.

The objective of this study is to determine the processes affecting the nature and magnitude of water storage changes during winter in a watershed containing wetlands, and the effect of these processes on the generation of streamflow. Winter streamflow is ecologically important to aquatic organisms, and it has economic implications in terms of culvert icing, and storage changes in winter affect basin water storage availability prior to snowmelt.

Study area

The study was carried out in a 17.4 km² drainage basin in Nipawin Provincial Park, central Saskatchewan (53°55′N, 104°45′W) (Fig. 1). Although the area is north of the mean annual 0°C isotherm, permafrost was not encountered in the study basin. At LaRonge, 125 km northwest of the area, the 30

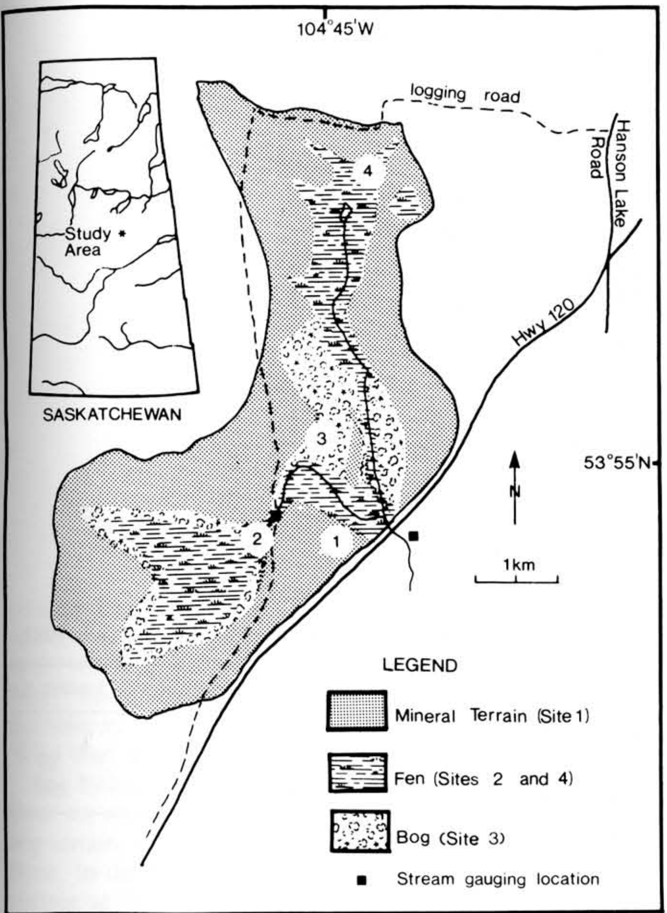

FIG. 1. Study area showing terrain types and instrumentation.

year mean annual precipitation is 485 mm, of which 164 mm is snow, and the mean January and July temperatures are −22.6 and 16.7°C respectively (Environment Canada 1982). The area has subarctic, mixed-wood vegetation, and mineral soils are primarily podzolic (Richards and Fung 1969).

The surficial geology is characterized by gently to moderately rolling till plain with a shallow layer (1.2−2.0 m) of sandy aeolian and fluvioglacial material overlying a compact clay till with low hydraulic conductivity (2.2×10^{-8} to 8.6×10^{-8} m/s).

The four terrain types found in the area are investigated in this study; mineral terrain, and wetland types including peat plateau bog, channel fen, and northern ribbed fen (Tarnocai 1979). These are described below. All four terrain types are represented in the 6.9 km² subcatchment in the southwest portion of the study area. Because of limited accessibility, however, all the study sites were not in the same basin. Mineral terrain constitutes 61% of the subcatchment, fen comprises about 34%, and bog 5%.

Site 1 is located on mineral terrain 250 m from the culvert on Highway 120, which drains the main watershed (Fig. 1). The sandy overburden is 1.2 m thick, with a vegetation cover of primarily white spruce (*Picea glauca*) and jack pine (*Pinus banksiana*), with some aspen (*Populus tremuloides*).

Site 2, herein referred to as lower basin fen, is located at the lowest end of the basin and transmits flow from the entire subcatchment. It can be classified as channel fen, occupying a small abandoned glacial spillway (Tarnocai 1979). Mean peat thickness here is 1.7 m, and is dominated by *Carex* sedges, with *Salix* (willow) and *Alnus* (alder) shrubs at the edges.

Site 3 is peat plateau bog, which has an irregular hummocky surface about 1 m above surrounding fen (Tarnocai 1979). It is located between the fen and the sloping mineral terrain, and has a locally perched water table. Subsurface drainage from the mineral terrain apparently passes beneath the bog, depriving it of nutrient-rich water. The mean peat thickness is 1.4 m, and the 60−80% hummocky ground cover comprises primarily *Sphagnum* mosses, with Labrador tea (*Ledum Groenlandicum*) on the hummocks and a 20−50% crown cover of stunted black spruce (*Picea mariana*).

Site 4, referred to as the upper basin fen, contains northern ribbed fen and horizontal fen, which are the most widely occurring organic terrain types in the area. This site is representative of the upper fen location in the gauged watershed. Peat thickness at site 4 is approximately 1.8 m, and the lowest lying central portion consists of alternating rib and trough (flark) microtopography oriented perpendicular to the direction of flow. *Carex* is the dominant surface cover, with alder on the ribs. Horizontal fen (Tarnocai 1979) is found nearer mineral terrain and is flat lying without a pronounced surface pattern, and has sparse tamarack tree cover (*Larix laricina*).

Methods

Climatic data

Maximum and minimum air temperature and continuously recorded air temperature measurements were made on the mineral terrain and at the bog sites. Supplemental data were obtained from the LaRonge, Saskatchewan, weather station. Temperature at LaRonge was strongly correlated with site data ($r^2 = 0.96$). Ground temperature was recorded manually at each site with a Soiltest meter and Soiltest moisture/temperature cells positioned at 0.1, 0.6, and 1.2 m below the surface.

Precipitation was recorded continuously at the mineral terrain site with a Universal type weighing gauge fitted with an Alter shield. A methyl-alcohol/ethylene-glycol mixture in the bucket liquified snow input and prevented freezing. Snow courses consisting of 10 graduated stakes were established in each of the terrain types to provide a spatially representative sample. Snow depth and density were measured on 30 March at 25 locations in each terrain type with a Mount Rose snow tube.

Ground freezing

To obtain depths of frost penetration a transect of 10 frost tubes was established at each site beside snow stakes. These were constructed according to the specifications of Rickard and Brown (1971). The 1.5 m long, 25 mm ID polyethylene inner tube was inserted into an immobilized outer tube (38 mm ID PVC) sealed at the bottom and placed vertically in the ground. The clear plastic inner tube contained sand saturated with a fluorescein solution (0.01%) which changed from green to clear when frozen. The length of the clear section was measured manually. The advantage of this method is that repetitive frost penetration measurements can be made easily and accurately at the same location. Rickard and Brown (1971) found the method accurate to within 20 mm.

Groundwater

Groundwater observation wells were located at sites chosen to be representative of each terrain type. At each site, a 0.1 m diameter perforated ABS pipe extended 1.5 m below the ground surface. The above-ground portion was unperforated, and was covered by an unheated but insulated housing. A

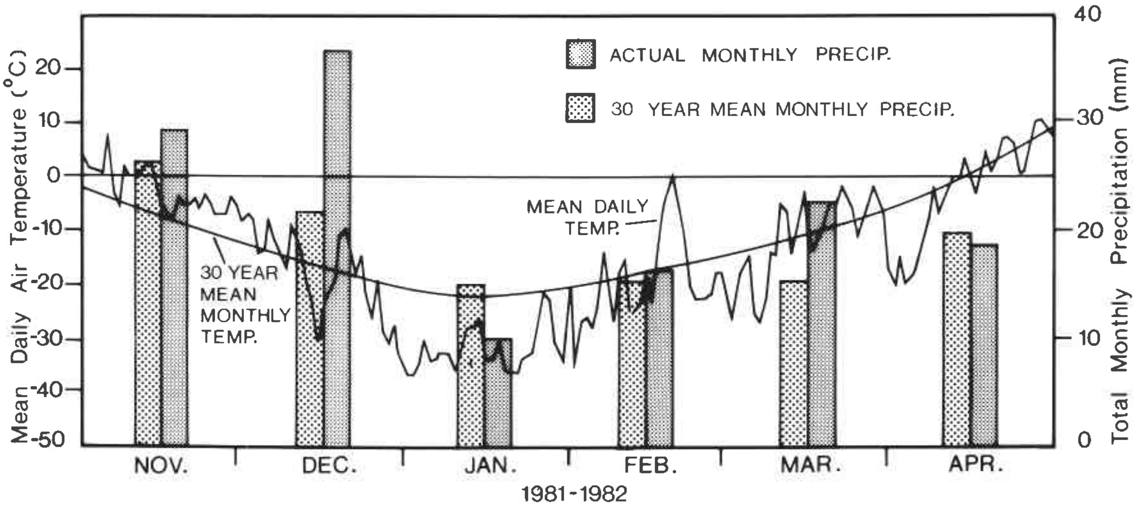

FIG. 2. Climatic normals and 1981 – 1982 data at LaRonge, Saskatchewan.

methyl-alcohol/ethylene-glycol solution was maintained in the tube to prevent freezing, and a light oil was added to prevent evaporation of the alcohol. A Leupold Stevens type F stage recorder continuously recorded the hydrostatic water level in each well. Hydrostatic water level in boreholes near to the well measured relative to the surface was within ±20 mm at the lower and upper fen sites respectively. The hydrostatic water level is defined as the level to which water would rise in a tube under atmospheric pressure.

Specific yield (S_y) was determined by gravimetric analysis on samples cored in the frozen state on 15 February 1982. The saturated samples of known weight and volume were allowed to drain for 24 h and the weight differential was measured to calculate specific yield. Saturated peat and mineral soil samples of known volume were also dried to calculate the total porosity, which is given as percentage water by volume of the undisturbed sample.

Streamflow

Stream velocity was measured with a Price type pygmy current meter at the outlet of the lower basin fen (site 2) where it crosses mineral terrain. Water at this location emerges from the fen less than 50 m upstream, and was sufficiently warm to prevent an ice cover from forming. A stage – discharge rating curve was used to estimate continuous discharge. Stage was recorded in a 0.3 m diameter PVC stilling well with a Belfort stage recorder. A similar installation was used on the main stream draining the main (17.4 km²) catchment downstream of the highway culvert (Fig. 1). However, an ice cover formed here in December and affected the stage readings.

Results

Climatic data

Mean daily temperatures of the winter of 1981/1982 were cooler than normal, and there was a heavier than normal December snowfall (Fig. 2). Mean daily temperature was below 0°C from 15 November 1981 to 10 April 1982, except for a day in February.

The last two rain events of the autumn season were 2.4 and 10.5 mm on 30 October and 15 November respectively. A permanent snow cover was established 21 November, with no significant melt until the following April. The average snow depth and density on 30 March are given for each terrain type in Table 1.

TABLE 1. Snow water equivalent and density on 30 March 1982

	Mineral	Bog	Upper fen	Lower fen
Depth (mm)*	94	107	97	89
Density (kg/m³)*	257	226	289	250

*White ice not included.

TABLE 2. Soil temperature (°C) at three depths in each terrain type

Depth	Mineral	Bog	Upper fen	Lower fen
17 October				
0.1 (m)	5.6	3.1	2.8	5.3
0.6 (m)	4.2	3.1	5.9	7.6
1.2 (m)	4.2	2.9	7.0	7.6
15 December				
0.1 (m)	−2.5	−3.9	−3.5	−2.9
0.6 (m)	0.5	1.5	1.0	3.3
1.2 (m)	1.2	2.1	—	4.0
15 February				
0.1 (m)	−8.5	−5.5	—	−2.7
0.6 (m)	−1.0	0.8	—	1.7
1.2 (m)	1.8	1.4	1.4	3.9

The mineral soils cooled more rapidly than the two fens (Table 2) because of the relatively low water content of the well drained sandy soil. The high water content of peat, which stores sensible and latent heat, resulted in higher temperatures at depth in organic terrains.

Groundwater

Groundwater movement in organic terrains is strongly affected by the two-layer wetland soil (Ivanov 1981). The peat characteristics measured from two cores per site are shown in Table 3. These properties typically change rapidly with depth in the acrotelm, but are less variable in the catotelm (Price and FitzGibbon 1982). The acrotelm thickness in bog and fen was

TABLE 3. Characteristics of mineral soil and double-layered organic soil

	Mineral*	Bog	Upper fen	Lower fen
Depth (m)[†]				
Acrotelm	0–1.20	0–0.12	0–0.20	0–0.15
Catotelm		0.12–1.70	0.20–1.40	0.15–1.80
Bulk density (kg/m³)				
Acrotelm	1400	47	76[‡]	99
Catotelm		64	96	130
Specific yield				
Acrotelm	0.28	0.24	0.31	0.29
Catotelm		0.15	0.21	0.13
Total porosity[§]				
Acrotelm	0.26	0.91	0.88	0.92
Catotelm		0.92	0.91	0.88

*Acrotelm/catotelm not applicable in mineral terrain.
[†]Zero is the surface elevation of the vertical profile.
[‡]Average bulk density, disregarding wood inclusion from a rib sample.
[§]Total porosity represents fraction of water by volume.

estimated to be 0.1 and 0.2 m respectively. The lower bulk density of the acrotelm (Table 3) indicates less decomposed peat, which corresponds to the higher specific yield found there. Thus, water can be more readily released from the acrotelm of the highly saturated peats (total porosity averages 90%) than it can from the catotelm.

The hydrostatic water level and the position of the freezing front are shown for each terrain type in Fig. 3. In mineral and bog terrain, the water table was consistently below the freezing front. In the fen sites, the water level was initially above the surface at the onset of freezing. When the surface water and peat froze, the hydrostatic water level continued to rise until mid-January. Thus, in the fens, the freezing front was always in direct contact with the saturated peat.

In mineral terrain, the water table was 0.4–0.8 m below the freezing front until mid-January, when it rose due to icing of the main stream at the culvert where it crosses the highway (Fig. 1). The hydrostatic water level of the stream rose to 0.45 m above the ice surface by 10 January, which affected the groundwater level at the nearby mineral terrain site. However, it was estimated by extrapolation (see Fig. 3) that the water table would have drained to the impermeable layer at 1.2 m by mid- to late January. Well drained sandy soil such as that on mineral terrains contains little unfrozen water behind the freezing front (Andersland and Anderson 1978); thus, it does not usually experience significant upward moisture migration in response to the temperature gradient (NRC 1984). The effect of such upward water migration on lowering the water table at this location is not known, but it is probably small.

In the bog, the water table decline indicated that most drainage occurred in November and December, when the gradient between it and the fen was greatest. Further drainage after this time may have been prevented by the high hydrostatic water level in the fen (see below). No ice lenses were found beneath hummocks, but small ice lenses were found in two of five cores taken from depressions at approximately 0.3 m. Freezing at this depth occurred after January; thus, it can be seen (Fig. 3) that it had little or no effect on lowering the water table at this location.

Both fens experienced positive storage changes early in the winter (Fig. 3) as a result of groundwater inflow. The hydrostatic water level rose between 16 November and 31 December (Fig. 4) because progressive freezing of the acrotelm (Fig. 5*a*–*d*) impeded drainage to the basin outlet. The rising hydrostatic water level displayed a stronger response during colder periods, and during a 22 mm (water equivalent) snowfall between 20 and 23 December, which may have caused compression of the acrotelm. During the period when the hydrostatic water level was rising, groundwater was observed emerging from the edges of a small (5 m × 3 m) spruce island, and along the stems of shrubs and bushes protruding through the frozen layer (see also Van Everdingen (1982)). White ice was formed as it mixed with the snow and refroze in a process analogous to the formation of a secondary white ice layer on lakes (Adams 1976*a*). White ice was formed up to the maximum hydrostatic water level, which was 91 and 47 mm above the 16 November level in the lower and upper fens respectively. The hydrostatic water level was sustained above the surface throughout the winter at the lower basin fen because its location at the lowest end of the watershed provided it with a large contributing area. At the upper basin fen, the hydrostatic water level dropped after mid-January because inflow from mineral terrain was diminished, and it has a relatively small area contributing to flow. However, the drop in hydrostatic water level did not cause desaturation because it did not fall below the freezing front. Therefore, the decline in hydrostatic water level in this confined aquifer represents the loss of a very small amount of water (Freeze and Cherry 1979, p. 61).

Streamflow

The last rainfall event of the year occurred on 15 November, producing a sharp rise in the basin hydrograph (Fig. 6). Recession to the prestorm streamflow rate was achieved after 72 h (18 November), but discharge continued to decrease until approximately 140 h after the rainfall (21 November). Recession to this level required 20 h/km², compared with 13 h/km² in a nearby wetland of similar size (FitzGibbon 1982), so the typical recession due to a precipitation event of this size was almost certainly over by 21 November. By 21 November, however, freezing had begun in the acrotelm (Fig. 5), thereby increasingly restricting flow to deeper peats of the catotelm and the underlying mineral soils. This caused another marked drop in streamflow (Fig. 6). From January onward, the frozen layer was within the catotelm (Fig. 5*e*–*g*), where increasing frost depth had much less effect on flow than it did as it penetrated the acrotelm. Streamflow during this period declined

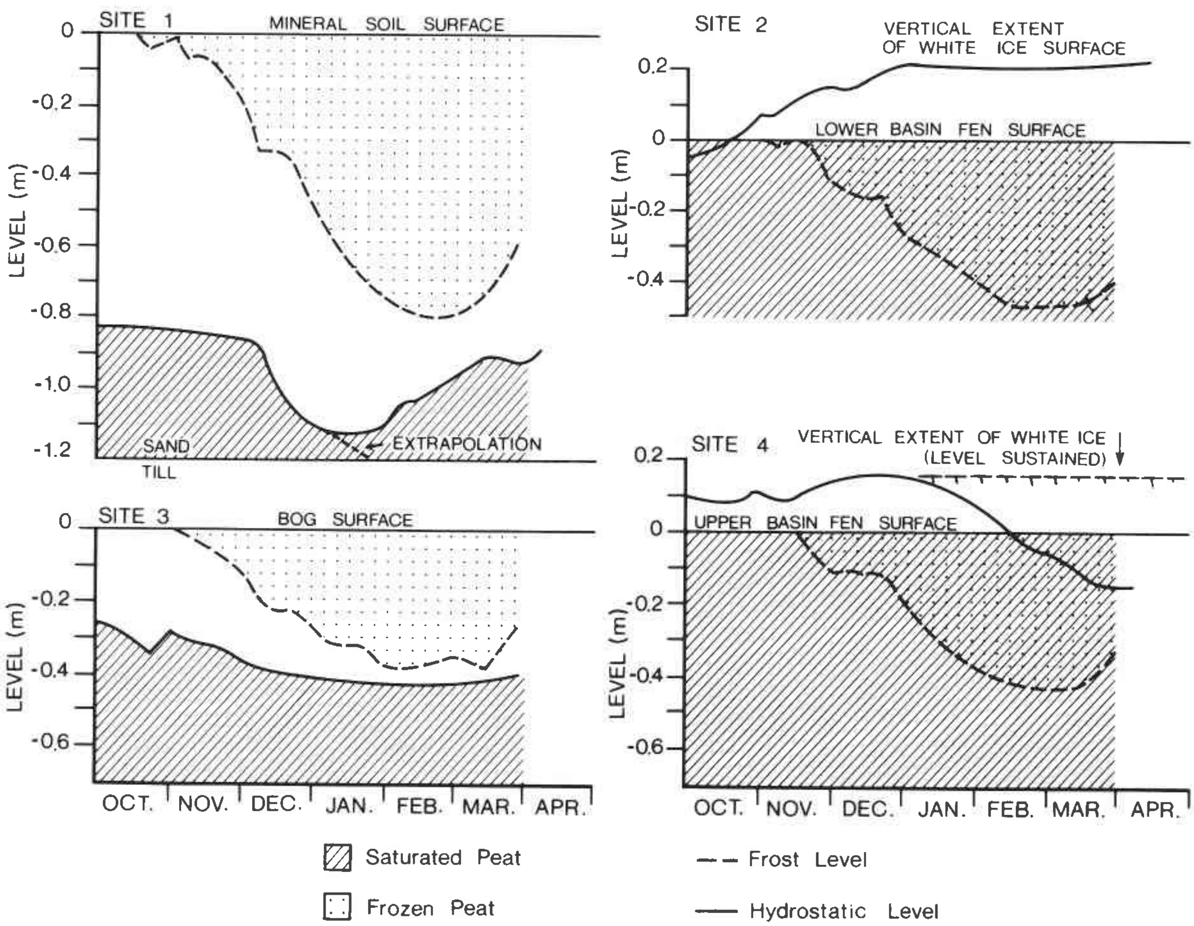

FIG. 3. Hydrostatic water level and frost depth with respect to the local surface. The extrapolation in mineral terrain indicates the probable water level if this location had not been affected by icing of a nearby stream.

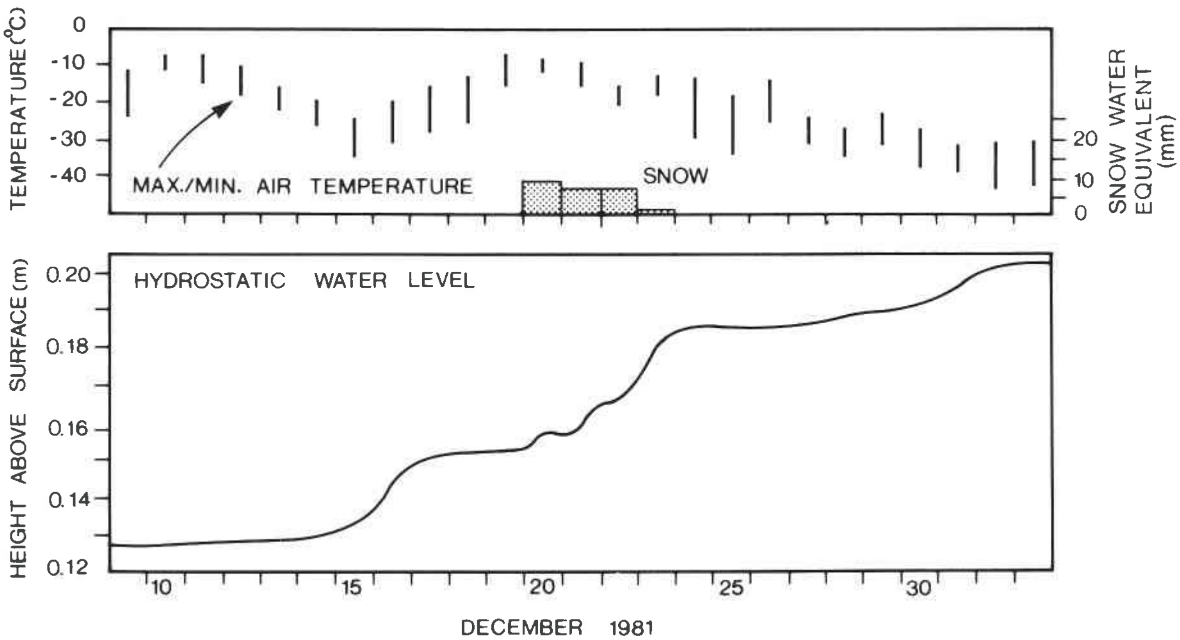

FIG. 4. Rising hydrograph of water table at the lower basin fen coinciding with periods of colder temperatures when the depth of frost in the acrotelm was increasing.

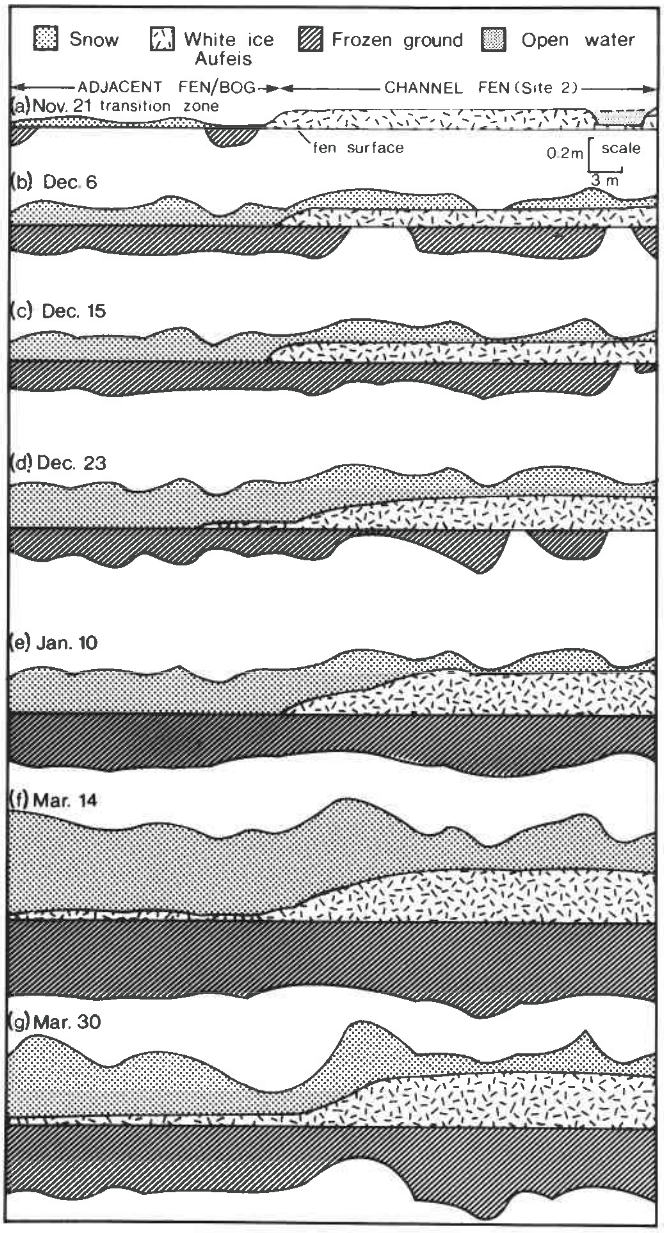

FIG. 5. Cross-sectional profiles across the lower basin fen and the adjacent transitional zone showing the extent of frost penetration, white ice growth, and snow at different times during the winter.

slowly and steadily from approximately 0.024 m^3/s to a pre-snowmelt discharge of 0.018 m^3/s, shown in Fig. 6 as the "winter baseflow" range. The basin average baseflow for the winter was approximately 0.25 mm/day, which totalled 35 mm between 16 November and 31 March.

Water balance

The water balance can provide a comparative overview of the various processes involved in wetland hydrology during winter. Unlike annual or summer water balances (Dooge 1972), winter precipitation can be neglected because it remains in storage as snow. Evaporation or sublimation losses from the snowpack can, therefore, be neglected, and the water-balance equation can be written

$$[1] \quad Q + \xi = \Delta S_m + \Delta S_b + \Delta S_f$$

TABLE 4. Storage changes (ΔS) at each terrain type based on the change in hydrostatic water level (Δh) and white ice thickness (ΔI) according to [2]

	Mineral	Bog	Upper fen	Lower fen
Δh (mm)	−335	−90	47*	91
ΔI (mm)	0	0	47	91
ΔS (mm)	−94	−25	28	51

*Represents only the positive change in height (i.e., that which caused white ice to form).

where Q is streamflow, ΔS is the change in storage, and subscripts m, b, and f refer to mineral, bog, and fen respectively. The error (ξ) is calculated as a residual.

Storage change (ΔS) at each terrain type is a function of (1) the change in hydrostatic water level (Δh) with respect to the specific yield (S_y) and (2) the white ice growth (ΔI) due to the refreezing of inflowing groundwater with snow at the surface. This can be calculated as

$$[2] \quad \Delta S = \Delta h \cdot S_y + \Delta I(\rho_i - \rho_s)$$

where snow density (ρ_s) (Table 1) is subtracted from the white ice density (ρ_i) to account for the contribution of snow to the white ice volume. A change in elevation of the hydrostatic water level (Δh) results from drainage of saturated soil to the unsaturated condition (or vice versa). It is assumed that there is no change in storage in the unfrozen zone unless desaturation occurs, and that the change of water from liquid to ice is not a storage change.

In mineral terrain and bog sites where the water table was consistently below the freezing front, $\Delta I = 0$. Change in storage in mineral and bog terrain, therefore, is

$$[3] \quad \{\Delta S_m, \Delta S_b\} = \Delta h \cdot S_y$$

In the fens, the hydrostatic water level was never lower than the freezing front, so desaturation did not occur; thus, $\Delta h \cdot S_y = 0$ and

$$[4] \quad \Delta S_f = \Delta I(\rho_i - \rho_s)$$

Accordingly, between 16 November and 31 March the change in storage of water in mineral and bog terrain was −94 and −25 mm respectively (Table 4). In the upper and lower fen, the storage change was 28 and 51 mm of water, respectively, assuming an average white ice density of 850 kg/m^3 (Adams 1976*b*).

The computed storage changes indicate that mineral terrain released almost four times as much water as the bog. Because of its large area, it is far more important in supplying water for streamflow, with sufficient excess to provide the storage gain experienced in the fens. The positive storage change in fens resulting from groundwater inflow from mineral terrain reduces the amount of water available for streamflow. Thus fens, rather than being a significant net contributor to streamflow during winter, simply transmit some of the water they receive, and store a significant portion.

It is useful to weight these storage changes according to their area and compare them with the basin streamflow. However, this must be done cautiously. Accurate evaluation of [1] depends on how representative the data are for each terrain type, which cannot be assessed given the limited number of observation wells. Further error can be introduced by the limi-

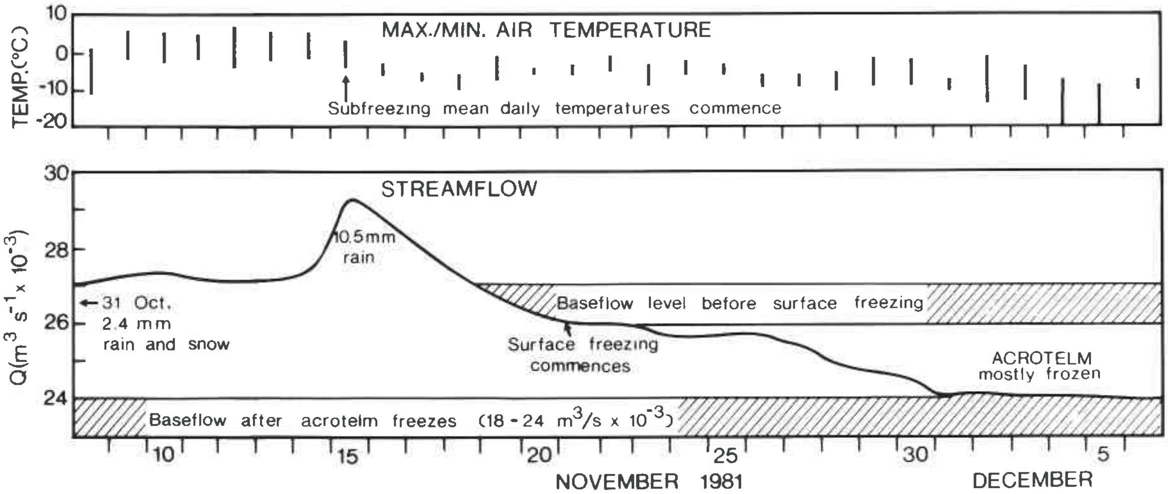

FIG. 6. Stream hydrograph during early winter showing diminishing streamflow corresponding to cold temperatures and increased frost in the acrotelm of the lower basin fen. The winter baseflow was achieved when the acrotelm was completely frozen.

TABLE 5. Water balance from 16 November to 31 March

	Mineral	Bog	Upper fen	Lower fen	Streamflow	Error
Percentage area	61	5	34		—	—
ΔS_a (mm)*	−57	−1	13		35	10

*Areally weighted change in storage.

tations of experimental techniques and measurement, e.g., ±15% for streamflow (Ackers *et al.* 1978) and ±12% for snow density (Goodison *et al.* 1981). The assumptions are that (1) the groundwater observation wells are representative of the terrain types in the subcatchment, (2) the hydrostatic water level in the mineral terrain site can be extrapolated as shown in Fig. 3, (3) moisture migration to the freezing front in the mineral terrain is negligible, and (4) the drop in hydrostatic water level in the confined aquifer of the fen is a negligible storage change. Thus, with the above caveat, [1] has been evaluated and the results are given in Table 5. The processes operating in the fens are similar, and the division between upper and lower basin fen area is arbitrary, so their storage changes were averaged. The average value is considered to be more representative of the entire fen than the values from the more extreme upper and lower fen locations.

Mineral terrain was the only significant net contributor (57 mm), whereas fens gained a much smaller amount (13 mm). The error calculated from [1] is 18% of the input by mineral terrain. Mineral terrain drainage far exceeded the volume of water lost as streamflow (35 mm); thus, it can account for the storage gain in the fens.

Discussion and conclusions

The wetland types discussed here are similar to many of those occurring throughout the subarctic, so these processes could be expected to occur elsewhere. The absence of permafrost permits water transfer between terrain types, resulting in considerable storage changes and streamflow during winter. However, the occurrence of permafrost would probably restrict this activity somewhat, depending on its extent.

In this study, mineral terrain and bogs were mostly unaffected by surface freezing. Groundwater in the saturated zone drained freely from storage, and was not recharged (except in the vicinity of the ice-blocked stream and culvert). Drainage of the bog catotelm was minor, which in view of its low permeability (Boelter 1965) and small areal extent in this basin resulted in little capacity to sustain streamflow. Similar processes probably resulted in Bay's (1969) observation that bogs produce no flow in winter.

Fens are fed by groundwater from mineral terrains, and occupy the valley bottoms. Continuous groundwater inflow maintained the water table at the surface, which froze in winter. When the acrotelm froze and flow through the system was constricted, the hydrostatic water level increased beneath the confining frozen layer. Groundwater seepage at the surface occurred and white ice was formed above the primary surface. Thus, water retained added significantly to the positive net storage (in fens).

The water balance indicates that the major water transfer during the winter was from the mineral terrain to the fen. Some of this was transmitted through the fen to produce streamflow, but much of it was retained as white ice. Therefore, rather than being instrumental in providing water for winter streamflow, the primary role of the fen was to transmit water that originated or was supplemented by recharge from mineral terrain. This contrasts with the findings of Bavina (1972), who also found that mineral terrain drainage provided inflow to fens, but that fens lost more water over winter than they gained. However, as no methodological details were provided, it is difficult to know what components were included.

A continuously flowing stream in an ice-free channel resulted in total basin storage loss of 35 mm, and net storage changes are reflected by the diminishing streamflow over winter. These changes control the available storage capacity in the basin at the time of spring melt. Mineral and bog terraines that have drained freely over the winter have their maximum available water storage capacity at the end of winter, whereas fens have very little.

Acknowledgment

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ACKERS, P., WHITE, W. R., PERKINS, J. A., and HARRISON, A. J. M. 1978. Weirs and flumes for flow measurements. John Wiley, Toronto, Ont.

ADAMS, W. P. 1976*a*. Diversity of lake ice cover and its implications. Musk-Ox, **18**: 86–99.

——— 1976*b*. Field determination of the density of lake ice sheets. Limnology and Oceanography, **21**: 602–608.

ANDERSLAND, O. B., and ANDERSON, D. M., *editor*. 1978. Geotechnical engineering for cold regions. McGraw-Hill, New York, NY.

BAVINA, L. G. 1972. Water balance of swamps in the European part of the USSR. Proceedings of the Minsk Symposium, UNESCO Press – IAHS, Paris, 1975, pp. 297–303.

BAY, R. R. 1969. Runoff from small peatland watersheds. Journal of Hydrology, **9**: 90–102.

BOELTER, D. H. 1965. Hydraulic conductivity of peats. Soil Science, **100**: 227–231.

DOOGE, J. 1972. The water balance of bogs and fens. Proceedings of the Minsk Symposium, UNESCO Press – IAHS, Paris, 1975, pp. 233–271.

ENVIRONMENT CANADA. 1982. Canadian climatic normals: temperature and precipitation, Prairie Provinces. Ottawa, Ont.

FITZGIBBON, J. E. 1982. The hydrologic response of a bog–fen complex to rainfall. Proceedings of the Canadian Hydrology Symposium, Fredericton, N.B., pp. 333–346.

FREEZE, R. A., and CHERRY, J. A. 1979. Groundwater. Prentice-Hall. Englewood Cliffs, NJ.

GOODISON, B. E., FERGUSON, H. L., and MCKAY, G. A. 1981. Measurement and data analysis. *In* Handbook of snow: principles, processes, management, and use. *Edited by* D. M. Gray and D. H. Male. Pergamon Press, Toronto, Ont., pp. 191–274.

INGRAM, H. A. P. 1978. Soil layers in mires: function and terminology. Journal of Soil Science, **29**: 224–227.

IVANOV, K. E. 1981. Water movement in mirelands. *Translated by* A. Thomson and H. A. P. Ingram. Academic Press, London, England.

NRC (NATIONAL RESEARCH COUNCIL). 1984. Ice segregation and frost heaving. National Academy Press, Washington, DC.

PRICE, J. S., and FITZGIBBON, J. E. 1982. Winter hydrology of a forested drainage basin. Proceedings of the Canadian Hydrology Symposium, Fredericton, N.B., pp. 347–348.

RICHARDS, J. H., and FUNG, K. I., *editors*. 1969. Atlas of Saskatchewan. University of Saskatchewan, Saskatoon, Sask.

RICKARD, W., and BROWN, J. 1971. The performance of a frost-tube for the determination of soil freezing and thawing depths. Soil Science, **113**: 149–154.

ROULET, N. T., and WOO, M. K. 1986. Hydrology of a wetland in the continuous permafrost region. Journal of Hydrology. (In press.)

SCHWARTZ, N. W., and MILNE-HOME, W. A. 1982. Watersheds in musket terrain. The chemistry of water systems. Journal of Hydrology, **57**: 267–290.

TARNOCAI, C. 1979. Canadian wetland registry. *In* Proceedings of a Workshop on Canadian Wetlands. *Edited by* C. D. A. Rubec and F. C. Pollet. Environment Canada, Lands Directorate, Ecological Land Classification Series, No. 12, pp. 9–21.

VAN EVERDINGEN, R. O. 1982. Management of groundwater discharge for the solution of icing problems in the Yukon. Proceedings of the 4th Canadian Permafrost Conference, Calgary, Alta., 1981. National Research Council of Canada, pp. 212–226.